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Heat Sources and Sinks at the Earth's Surface

by

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(Meteorological Office)

AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

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Summary

A study has been made of the spatial and temporal variations in the surface character of the earth which can result in increases or decreases in the heat supply to the atmosphere. For the sea areas the changes in surface temperature and in the ice limits are considered and for the land the changes in albedo, evaporation and evapotranspiration, state of surface, heat-storage capacity and conductivity are dealt with. Some of these factors are found to be unimportant, while others can exert an appreciable and sometimes long-term influence upon the heat supply to the atmosphere.

1. INTRODUCTION

The solar radiation received at the outside of the atmosphere is 800 to 900 cal cm^{-2} day^{-1} throughout the year at the equator and at the poles it is 1070 cal cm^{-2} day^{-1} in midsummer, falling to nil in winter. If we take a long period average covering the northern hemisphere then approximately 47 per cent. of this radiation reaches the earth's surface and about 19 per cent. is directly absorbed by the atmosphere and clouds (Houghton, 1954). Some of the radiation received at the earth's surface is reflected back to space, some is absorbed and stored and some is passed to the atmosphere by convection and radiation. All of these processes are affected by the physical character of the surface, with the result that the amount of heat reaching the atmosphere either immediately or by the subsequent transfer of stored heat from the surface may be significantly affected. Here we are concerned with those effects which may be sufficient to affect the circulation of the atmosphere over a substantial area for a period of days. Such effects must occur over a substantial area and persist for a few days at least. We therefore limit the study to changes which can affect an area of a minimum of 10^4 Km^2 for 5 days at least.

The magnitudes of the variations in heat supply to the atmosphere may be compared with the outgoing long wave radiation to space (500 cal cm^{-2} day^{-1} in low latitudes, 380 cal cm^{-2} day^{-1} in high latitudes), or with the change in the heat reflected to space (and consequently lost to the atmosphere) due to a change in cloudiness from nil to 8/8 (100 to 200 cal cm^{-2} day^{-1} in low latitudes). However, it is probably safer to compare the variations with one another and see which are the most important.

The energy of the atmospheric circulation is derived primarily from horizontal differences of temperature. These horizontal temperature gradients

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may be affected by variation in the supply of either sensible or latent heat to the atmosphere. There is however the important difference that the modification to the temperature field resulting from variations in supply of latent heat will take place when condensation occurs and this may be at a distance of several hundred kilometres from the source of evaporation.

2. THE OCEANS

Of all the heat stored and released by the earth's surface during the course of the annual temperature cycle 86 per cent. is due to the oceans (Schumann and Burdecki (1953)). Most of this is released in winter months and at a rate which depends upon the temperature difference between the sea and the air above. As far as the sea itself is concerned the determining factors in its heat output are its surface temperature and whether or not it is frozen.

2(a) Sea surface temperature

When the sea surface is warmer than the air above it, heat is transferred upwards by convection and long wave radiation. Convective processes are also set up in the sea itself so that the sea surface temperature changes little.

When the sea surface is colder than the air above it, stable conditions are established in both media and heat is transferred downwards by long wave radiation and turbulent diffusion. Cloudy conditions are likely and there will be net downward radiation from the cloud base, but even for a cloud-sea temperature difference of 5°C (which is rare) this is only about 25 cal cm^{-2} day^{-1}.

A relationship between the flux of heat and water vapour to the atmosphere and the air-sea temperature difference has been suggested by Jacobs (1942) but is applicable only to mean values over a period of about a season. Special cases covering shorter periods have also been analysed from time to time and provide some indication of the variation in intensity of convective heat transfer over the oceans.

(i) The long term heat budget

Jacobs gives the empirical relation

$$Q_e = 0.143\ (e_w - e_a)W_a$$

where Q_e = heat equivalent of evaporated water (cal)

W_a = wind speed ($m.sec^{-1}$)

e_w = saturation vapour pressure at sea surface temperature (mb)

e_a = vapour pressure of the air at deck level (mb).

The associated transfer of sensible heat (Q_c) is then found using Bowen's ratio. No difference is made between rough and smooth conditions.

Jacobs' charts showing the mean seasonal values of Q_c and Q_e over the N. Pacific and N. Atlantic Oceans give winter values for Q_c of 20 to 80 cal cm^{-2} day^{-1} to the west of Ireland rising to over 200 cal cm^{-2} day^{-1} off the east coast of North America. The annual means for Q_c are 20 and 100 cal cm^{-2} day^{-1} respectively. Corresponding values for Q_e are about $2\frac{1}{2}$ times Q_c and the annual means for ($Q_c + Q_e$) rise from about 70 cal cm^{-2} day^{-1} off western Ireland to 350 cal cm^{-2} day^{-1} off eastern America.

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The observed annual amplitudes of sea surface temperature at the weather ship stations 'I' and 'J' support these figures. The mean amplitude through the top 100 metres is about 4.5°C, so that the heat given up by the sea over the 6 winter months is about 45,000 cal cm^{-2}, an average of about 250 cal cm^{-2} day^{-1}. Of this about half is accounted for by long wave radiation. Sverdrup (1940) quotes similar figures for the Bay of Biscay. Over these areas of the eastern Atlantic the annual mean of the air-sea temperature difference is between -1°C and -2°C.

(ii) Shallow convection

Riehl (1951) has studied the atmospheric heat balance over a period of a few days in the trade wind regions of the North Pacific Ocean. His data satisfy the empirical equation $Q_c = 0.036\ (T_w - T_a)\ W_a$, with values of Q_c ranging from 10 to 30 cal cm^{-2} day^{-1} for an air-sea temperature difference of 0.5 to 1.0°C.

Data for the Baltic are contained in a paper by Hela (1951) who studied the heat exchange at Finngrundet (a lightship). He gives the mean input of sensible heat to the atmosphere during winter months as about 50 cal cm^{-2} day^{-1}. It is probable that shallow convection obtained on the majority of occasions, so that this figure is applicable here. He suggests the equation $Q_c = 0.032\ (W_a + 0.3)^{\frac{1}{2}}\ (T_w - T_a)\ \Theta_w$ to fit his data, where Θ_w is the sea surface temperature in degrees absolute.

For cases of shallow convection over the North Sea but with a much larger temperature difference (about 10°C) the author has found the sensible heat input to be about 40 to 80 cal cm^{-2} day^{-1}. The data used were for very cold easterlies circulating around a Scandinavian anticyclone.

(iii) Strong convection

Craddock's (1951) paper on the modification of polar airstreams contains useful data. He shows that the sensible heat input to the atmosphere can reach values as high as 1400 cal cm^{-2} day^{-1}, while 300-500 cal cm^{-2} day^{-1} is probably quite normal in direct airstreams from the north. These figures are not strictly comparable with those already given, as an unknown proportion of the observed heating is due to release of latent heat in the shower precipitation during the passage of the air over the sea. The equations $Q_c = 0.064\ (T_w - T_a)$ and $Q_e = 0.017\ L\ (T_w - T_a)$ fit his graphs relating the sensible heat and water vapour transfer to the air-sea temperature difference.

A similar type of air mass is studied by Burbidge (1951). Using his data for the modification of continental polar air crossing the Hudson Bay in the late autumn the input of sensible heat is at least 344 cal cm^{-2} day^{-1}.

No account has been taken by Craddock or Burbidge of the long wave radiational cooling of the atmosphere for which the relevant figure is of the order of 150 cal cm^{-2} day^{-1}.

Winston (1954) relates intense cyclogenesis in the Gulf of Alaska with areas of rapid heating from the sea surface. He finds the non-adiabatic component of heating in the five cases of cold air masses studied to lie between 400 and 2210 cal cm^{-2} day^{-1}. Unfortunately values of sea and air temperature are not given.

(iv) Estimation of the effect of an anomaly of 1°C in sea temperature on the heat transport to the atmosphere

Several relationships have been suggested which give some indication of the dependence of the supply of sensible and latent heat to the atmosphere upon the sea surface temperature. For the mean effect over a period of months the partly empirical equations derived by Jacobs and Hela indicate that an

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anomaly of ± 1°C in air-sea temperature difference will give ± 30 cal cm^{-2} day^{-1} in sensible heat, and ± 60 cal cm^{-2} day^{-1} in latent heat available to the atmosphere. A relationship based on Craddock's analysis suggests values up to ± 80 cal cm^{-2} day^{-1} for the increase in the supply of sensible heat corresponding to ± 1°C in strongly convective situations, while the latent heat supply will change by ± 160 cal cm^{-2} day^{-1}.

Now a change in the temperature of the sea surface is always accompanied by a change in the temperature of the air immediately above it. It is reasonable to suppose that in the case of off-shore advection of comparable air masses over sea areas close to continents an anomaly of 1°C in sea surface temperature will correspond to an anomaly of 1°C in air-sea temperature difference. The above figures can then be applied as they stand and they may be assumed to indicate the maximum effect on the heat input to the atmosphere of a 1°C sea temperature anomaly. In all other cases the air-sea temperature difference anomaly will be some fraction of the sea temperature anomaly. An independent estimate of the size of this fraction may be obtained by considering the change in heat storage capacity of the sea itself. Anomalies in sea surface temperature are sometimes observed to last for several months. If, as is reasonable, a positive anomaly of 1°C extends to 50 metres depth and disappears after 5 months then the atmosphere has received an additional 50 cal cm^{-2} day^{-1} (corrected for approximately 5 cal cm^{-2} day^{-1} lost as additional long wave radiation). This suggests that the mean effect of a 1°C anomaly is appreciable over quite a long period and somewhere near half the maximum effect indicated by the empirical equations.

2(b) Ice

(i) Freezing

When the sea freezes the transfer of heat from the sea to the air is very much reduced. This is demonstrated most markedly in observations of the modification of polar continental air over Hudson Bay before and after freezing (Burbidge, 1951). The daily input of heat to the air crossing the Bay before the surface freezes is at least 344 cal cm^{-2} (see para. 2(a)). After the Bay freezes in December hardly any modification can be observed.

(ii) Conduction

If the temperature of the water underlying the ice is 0°C and the air temperature well below this value, the rate of heat transport through the ice may be estimated for various thicknesses of ice. For ice 1 metre thick and a temperature difference across the ice of 10°C, assuming a thermal conductivity of 0.005 cal $(sec\ cm\ °C)^{-1}$, it is about 20 cal cm^{-2} day^{-1}. In areas where the sea freezes the atmosphere loses heat by long wave radiation at the rate of about 380 cal cm^{-2} day^{-1}. Hence any heat received by conduction through the ice is negligible in comparison with the loss of heat by radiation.

(iii) Albedo

Changes in the albedo of the sea through freezing need not be considered, as the incident solar radiation in the areas concerned is either nil or too small to matter.

2(c) Ocean Currents

The subject of ocean currents is beyond the scope of this paper, but it must be noted that it is these currents which largely determine the latitudinal and longitudinal distribution of sea temperatures. A sea temperature anomaly is normally due to either an anomaly in absorbed radiation or an anomaly in the direction or rate of flow of ocean currents. Also, as a result of ocean currents the areas of anomalous radiation and resulting anomalous sea temperature may become far removed from one another. The time lapse is important. For instance a large sea temperature anomaly in the Caribbean Sea and off the coast of Florida

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may be expected to be transferred northeastwards in the Gulf Stream and to reach the northeastern Atlantic in the course of a year. The maximum flow which is found off Chesapeake Bay is about 3 knots, while over most of its course the Gulf Stream transports water at about 1/3 to $\frac{1}{2}$ knot (Sverdrup, 1940).

2(d) Other aspects

Other aspects of energy transformation over and in the oceans may be mentioned but are unimportant. Flow of heat through the ocean bottom is about 0·18 cal cm^{-2} day^{-1} (Helland-Hansen, 1930). Frictional energy of the wind on the sea surface is estimated at 10^{-4} times the radiation received at the surface (Sverdrup, 1940). Tidal dissipation of energy over shallow coastal waters is about 2·9 cal cm^{-2} day-1 (Taylor, 1919).

2(e) Heat exchange between atmosphere and ocean - summary

The exchange of heat (sensible and latent) between the sea and the atmosphere is negligible when the sea is frozen, and small ($\sim$20 cal cm^{-2} day^{-1}) when it is colder than the air. It is significant ($\sim$ 50 cal cm^{-2} day^{-1}) when mechanical turbulence and shallow convection pass heat to the air and it is very large (up to 1200 cal cm^{-2} day-1) in strongly convective situations. Downward long wave radiation may supplement the small turbulent heat flux when the sea is colder than the air but is normally much smaller than 25 cal cm^{-2} day^{-1}.

The maximum change in heat supply to the atmosphere over the open ocean resulting from an anomaly of ± 1°C in sea surface temperature is estimated at ± 15 cal cm^{-2} day^{-1} for sensible heat and ± 45 cal cm^{-2} day^{-1} when latent heat is included. This applies to conditions averaged over a period of months. The latter figure is confirmed by an estimate based upon maximum observed sea temperature changes. In strongly convective situations these values may be increased to ± 40 and ± 120 cal cm^{-2} day^{-1} respectively.

Observed sea temperature anomalies of the order 1 to 3°C and the corresponding increase or decrease in energy made available to the atmosphere must have some effect on subsequent synoptic developments.

The freezing of the sea surface, although it has been shown to be important, only occurs on a comparatively small scale. The freezing of large areas of partially enclosed seas which is an annual occurrence in high latitudes is most important in connection with the modification of local air masses. It is possible that the late freezing of areas such as Hudson Bay might also hinder the establishment of the semipermanent continental cold troughs in the troposphere, and the abnormal freezing of any such area should also have an observable effect on the subsequent thickness pattern in the troposphere. However there are few areas in which the dates of freezing of substantial areas of the sea have significant variations. For instance the Baltic Sea is subject to a certain amount of freezing every winter, but it seems improbable that a variation of 20 to 50 miles in the freezing limit can exert any appreciable large scale control on the atmospheric circulation.

3. THE CONTINENTS

The non-uniformity of the land surface and its comparative inability to store and conduct heat necessitate a different approach to the question of whether any changes in its character can appreciably change the heat supply to the atmosphere.

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Little of the heat which is received by a land surface from the sun is stored in the ground for more than 24 hours. Some of it is returned to space by reflection, and radiation, some passes to the atmosphere by long wave radiation or convection and some is used in evaporation in the melting of snow etc. Any increase in the amount of heat reflected or in the amount used in evaporation will reduce the amount of sensible heat made available to the atmosphere and it is therefore convenient to think of the increase of heat dissipation by any such process as a potential local loss to the atmosphere.

3(a) Heat conduction and storage

(i) Unfrozen ground

Heat flow to and from the surface depends upon the thermal conductivity of the materials in the surface layers, on the seepage of water downwards associated with rainfall and on the temperature gradients which are established primarily as a result of incoming and outgoing radiation. All these are very variable but the purpose of this study will be achieved if extreme values can be estimated.

Using known values of thermometric conductivity (K) density (ρ) and thermal capacity (c), the following values of heat flux are calculated assuming a temperature gradient of 10°C per metre.

Table 1

	ρc	$K \times 10^{-2}$	Flux cal cm^{-2} day^{-1}
Wet sand	0.53	0.60	28
Moist sand	0.34	0.56	16
Dry sand	0.30	0.16	4
Sandy soil	0.35	0.62	19
Dry loam	0.4	0.2	7
Marsh	1.00	0.12	10
Light soil with roots	0.1	0.3	3
Uncultivated soil	0.52	0.34	15
Clay	1.4	0.2	24
Rock	0.5	1.2	52
Snow	0.12	0.2	2
Ice	0.45	0.78	30

Temperature gradients of the order of 10°C per metre are only found in conditions of strong insolation, or radiation under clear sky at night, but for short periods they may be considerably greater in the top few centimetres. On a hot summer's day at Dresden, Schreiber (1912) found the maximum mid-day heat flux into the sandy soil to be about 70 cal cm^{-2} day^{-1}. Roach (1955) using a flux plate at 6 cm depth at Kew measured a maximum flux of 5 mw (100 cal cm^{-2} day^{-1}) which lasted for about 3 hours on a hot June day. With the

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same instrument he measured a maximum upward flux of 1 mw during a very cold easterly outbreak in January, 1954. This flux persisted for at least 24 hours. Confirmatory evidence is obtained by examining earth temperature records for Berlin and Dublin which show that the maximum 24-hour temperature changes recorded with warm air flowing over cold land or vice versa are about 3°C in the top half metre which correspond to a maximum heat flux of 30 cal cm^{-2} day^{-1}. For a long period average the total change in the heat content of the soil at Berlin between October and March gives a value of about 1700 cal cm^{-2}. This may be compared with the corresponding figure for the eastern North Atlantic which is about 45,000 cal cm^{-2} (see para. 2(a)). Thus we see that the modification of air masses by sensible heat stored in the ground is not an important process in the general circulation.

(ii) Frozen ground

Normally the ground is moist before winter freezing takes place and the soil is maintained for a period at 0°C while heat is abstracted from or communicated to it during freezing or thawing. Higher temperature gradients than are normally observed may thus be established above a frozen layer of soil and may persist for considerable periods.

Let us consider a square centimetre of earth 10 cm in depth which contains 4 gm of liquid water. The heat liberated during the freezing process is about 320 calories and the normal time taken to freeze to this depth is 4 to 10 days, giving a maximum heat flux over 4 days of 80 cal cm^{-2} day^{-1}. Thawing may proceed rather more rapidly as solar heating of the soil surface gives higher temperature gradients. At Berlin 10 cm are observed to thaw in as little as 2 days and 25 cm in 7 days. The heat absorbed is a maximum of about 150 cal cm^{-2} day^{-1}, some of which is directly removed from the atmosphere and the rest would otherwise have been transmitted to the atmosphere by convection from the heated ground. Thawing by the advection of a warm cloudy airstream proceeds at about half this rate and the maximum heat lost from the atmosphere by radiation and diffusion is about 80 cal cm^{-2} day^{-1}.

These fairly high values of heat flux are important because of the length of time for which they may persist. They suggest that frozen ground could prevent normal atmospheric heating to a significant extent, so that variations in the semi-permanent limits of frozen ground could be important.

3(b) Radiative properties

(i) Albedo

Reasonably accurate measurements of albedo are available for most types of vegetation and bare earth surfaces, though the dryness of the vegetation or earth can give quite wide variations. The following values are to be found in the literature (Budyko (1954), Fritz (1948), Houghton (1954), Smithsonian Meteorological Tables, Sutton (1953), Sverdrup (1940)).

Table 2

Percentage of incident radiation which is reflected

Forest	3 - 10
Snow covered forest	10 - 25
Grass fields dry	20 - 30
Grass fields wet	8 - 20
Bushes	10
Bare ground dry	10 - 25

Bare ground wet	8 - 10
Sand	20 - 25
Rock	12 - 15
Snow new	70 - 80
Snow old	55
Water	5 - 50 (depending on solar elevation)
Desert	24 - 30

In order to estimate the maximum effect of a change in albedo on the heat supply to the atmosphere it is reasonable to assume that all the radiation absorbed at the ground is redistributed to the atmosphere during 24 hours.

The largest albedo change is from a normal grass or earth surface (15 to 20 per cent.) to a snow covered surface (70 to 80 per cent.). Much winter snowfall occurs when the incoming solar radiation is so small that a change in albedo is insignificant in its effect on the heat supply to the atmosphere. However some areas as far south as 50°N, notably the southern shores of Hudson Bay (Thomson (1954)) and parts of central Siberia receive their first snowfall towards the end of September, when the mean sun and sky radiation is 180 cal cm^{-2} day^{-1} (Black (1956)). In this case the 60 per cent. increase in albedo consequent on snowfall will reduce the heat supplied to the atmosphere by 108 cal cm^{-2} day^{-1}. These same areas are the last to lose their snow cover in such southern latitudes and substantial areas of snow covered ground may persist until late April when the mean incident sun and sky radiation is as much as 300 cal cm^{-2} day^{-1}; 180 cal cm^{-2} day^{-1} is then denied to the atmosphere. Between these extremes the importance of snow cover decreases with increasing latitude and increasing proximity to the winter solstice. Snow cover in Scandinavia for instance will only be important before the end of October or after the end of February.

Of other possible changes in albedo the most important is from 15 per cent. with green grass to 25 per cent. with parched earth. This may occur in areas where the incident radiation is 500 cal cm^{-2} day^{-1}, when the 10 per cent. change in surface reflectivity will increase or decrease the heat available to the atmosphere by 50 cal cm^{-2} day^{-1}.

3(c) Snow

The change in albedo due to snowfall has already been discussed. There remains the effect of the absorption of latent heat on melting. A maximum rate of melting is about 5 cm of snow per day (about 0·5 gm) and the corresponding absorption of latent heat is 40 cal cm^{-2}. As in the case of thawing earth this is a loss to the atmosphere.

3(d) Evaporation and evapotranspiration

The thermal effects on a land surface as a result of evaporation and evapotranspiration are substantially different from the corresponding effects on a sea surface. By means of convection in the sea surface the temperature remains almost constant throughout the evaporation process. Over land the latent heat of evaporation must be supplied either by the sun, by the atmosphere or by conduction through the surface. As we have seen, the heat made available by conduction from that stored in the soil is negligible and the heat required for evaporation can be regarded either as a direct loss to the atmosphere

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or deducted from the solar energy which would otherwise have been communicated to the air by convection.

Numerous measurements indicate that the normal daily evaporation varies from 0·6 cm per day from a moist surface in the tropics to 0·05 cm per day from a dry surface in high latitudes. Table 3 gives a selection of values from the literature for evaporation and evapotranspiration.

Table 3

Surface and vegetation	Evaporation cm per day	Source
Grass-Toronto-November	0.05	Mather (1954)
Very dry soil - California	0.06	Veihmeyer (1954)
Bare soil - Toronto - November	0.07	Mather (1954)
Dry soil with vegetation - Florida	0.2	"
Bahia grass - Belgian Congo	0.3	"
Bare soil - Toronto - July	0.4	"
Moist soil with vegetation - Florida	0.5	"
Grass - Toronto - July	0.5	"
Moist sand - California	0.63	"
Grass - British Isles (Annual Mean)	0.14	Penman (1949)
	cm/year	
Savanna - tropics	92.5	Mather (1954)
Sugar cane	124.5	"
Very wet rain forest	152.9	"
Sawgrass	171.5	"

If all the latent heat of evaporation is considered to be a loss for the atmosphere then the deficit can be as much as 260 cal cm^{-2} day^{-1}. This deficit will vary according to the vegetation or crops growing and the moisture available. After a prolonged drought the heat available to the atmosphere may be as much as 200 cal cm^{-2} day^{-1} more than after a prolonged wet spell. This is partly offset by the increase in the albedo; but even so the net result should be evident in a much greater heating of the atmosphere over deserts in sub-tropical regions than over areas covered in green vegetation. The temporary formation of a new desert or vegetation over a normally desert area should be significant.

It has been assumed in the foregoing paragraphs that the heat used in evaporation is not returned to the atmosphere by recondensation in the same region. If showers occur this will not be true nor if the evaporation is balanced by deposition of dew. However in many circumstances the air will be transported hundreds of kilometres before the latent heat is released in some region of generally ascending air such as the frontal system of a depression. Under these circumstances any variation in the evaporation in a particular area may be regarded as providing an abnormal heat sink for the

/atmosphere

atmosphere in the area of evaporation and a corresponding abnormal heat source in some other region. The net world wide effect over a period will always be nil.

3(c) Summary

The probable maximum effects of variations in land surface conditions on the heat supply to the atmosphere are as follows:-

Table 4

Method	Source	Magnitude cal cm^{-2} day^{-1}
Conduction	Calculated - wet sand; temp-grad 10°C per metre	± 28
	Observed - Berlin; glacial ~~and~~ sand and gravel	± 30
Freezing	Latent heat liberated: calculated: water content 0.4 gm per c.c.	± 80
Thawing ground (sunny conditions)	Latent heat absorbed: same water content	- 150
Thawing ground (cloudy conditions)	"	- 80
Thawing snow	Latent heat absorbed in melting 5 cm per day	- 40
Albedo	Change due to snow: middle latitudes late autumn and early spring	- 180
	Change due to drought on green vegetation	- 50
Evaporation	Possible change due to drought or vegetation on desert	± 200

4. CONCLUSION

Before reviewing the ability of the earth's surface to modify through various means the working of the atmospheric heat engine, mention must be made of the solar constant. For many years it has been suggested that variations in the solar constant can give rise to large scale climatic changes and that small variations due to sunspot cycles are responsible for shorter fluctuations in world weather. Krivsky (1954) purports to find a definite relationship between sunspot cycles and the position of the jet stream. The most useful summary of the present state of this subject is by Wexler (1953). Although he is probably more disposed than many to advocate a relationship between sunspot cycles and the subsequent weather he states that there is as yet no proof of one and, what is more, that no proof can be readily produced. Correlations between random factors have been found which are equal to those found between weather phenomena and sunspot cycles.

The foregoing survey has covered the principal ways in which variations in the state of the earth's surface can change the energy available to the atmosphere. The following appear to be the most important:-

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(i) An anomaly in sea surface temperature.

(ii) Variations in the limits of frozen land and sea.

(iii) Early and late snowfall in middle latitudes.

(iv) Change from moist land or land covered in green vegetation to desert and vice versa.

A natural sequel to the present investigation is to seek evidence in the behaviour of the atmosphere of the effect on subsequent weather of anomalies of surface conditions of the kind listed above. The effects of the limits of arctic ice have already been considered by Schell (1940) and Walker (1947) but the relations with subsequent weather have been found to be weak. Possibly this might be expected from the very limited variations in the boundary of arctic ice.

A summary of other attempts to find associations between sea surface temperature, sea ice anomalies and subsequent weather has been made by Baur (1951). In particular Sandstrom suggested that a warm Gulf Stream should result in a warm winter over Europe. However the winter which followed a peak in the Gulf Stream in 1939 was the coldest in Europe for 110 years and the above-normal sea temperatures of 1955 were also followed by an abnormally cold February. Thus the search for such relations will not be easy and the effect of anomalous heat sources and sinks may be found more easily in anomalies of the atmospheric circulation over a wide area rather than in direct effects on local temperature.

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